IMPACT PRODUCED CONDENSATE AND DROPLET SIZE DISTRIBUTIONS

John D. O'Keefe and Thomas J. Ahrens

Seismological Laboratory
California Institute of Technology
Pasadena, CA 91125

March 7, 1989

(NASA-CR-186750) IMPACT PRODUCED CONDENSATE AND DROPLET SIZE DISTRIBUTIONS (California Inst. of Tech.) 35 p N90-71422

Unclas
00/70 0292221

Submitted to *Icarus*

ABSTRACT

The physics of condensation of impact produced vapor is described for a range of impactor radii from 1 to 10^7 cm and impact velocities from 15 to 45 km/s. We consider silicate projectiles impacting silicate bodies and ice projectiles impacting ice bodies. The height above the impact point at which condensation occurs was calculated. Also, we calculated the mean radius of the condensate and the distribution of condensate particle sizes. The altitude that condensation of shock-induced vapor for ice and silicate targets occurs is on the order of $\sim$10 times the projectile radius. The mean radius of droplets condensed at these altitudes, for silicate or ice impactors having radii from 1 to 10 km is in the range of $\sim$1 to 10 cm. Only some 10^{-4} times the mass of the impactor is found to condense into droplets with dimensions of radius μm or less. For the proposed Cretaceous-Tertiary extinction bolide, this implies that some 10^{14} to 10^{15} g of submicron-sized material would be generated. This is less than (10^{16}g) which is proposed to be required for the decreases in solar insolation leading to world-wide climatic changes.

INTRODUCTION

The meteorite impact velocities that are typical of the latter stages of planetary accretion, at the present time, are sufficient to produce vaporization on terrestrial and Saturnian planets and the icy satellites. The impact-induced vapor upon expansion from the impact site will condense into solid particles and the melt entrained in the expanding vapor will be broken-up into small droplets. The distribution of these particles can play a significant role in the thermal regime of the planet and satellite accretion, especially in the presence of an atmosphere as impact produced aerosols strongly affect the radiative balance during planetary accretion (Matsui and Abe 1986; Kaula, 1979). In addition, the distribution of aerosol-sized ejecta from a major impact such as that proposed at the Cretaceous/Tertiary boundary can give rise to an epic of massive extinction of biota.

The discovery in 1980 of a layer of platinum group elements (PGE), Ir, Os, Au, Pt, Re, Ru, and Pd-rich clay at the Cretaceous-Tertiary boundary (65 Ma) in two shallow marine localities in Europe has raised considerable debate as to its origin and its relation to the long observed major biological extinctions (Alvarez et al., 1980). Alvarez et al. (1980) hypothesized that the ejecta from a very large impact on the earth would be globally distributed in the stratosphere, and be deposited world-wide in a cm thick layer of a Pt element enriched layer. During the period that aerosol-sized ejecta ($<1\mu m$) is suspended in the atmosphere, it has been proposed that the decrease in solar insolation gave rise to a sudden global cooling event which triggered the long recognized mass extinctions at the Cretaceous-Tertiary boundary. Critical to this scenario is an understanding of the distribution of impact induced aerosols from gas, liquid, and solid ejecta.

Finally, the discovery of microtektite sized spheres also at the K-T boundary (Smit and

Hartogen, 1981; Kyte et al., 1985) and the recent discovery of the association of microtektites, tektite debris, and shock metamorphosed minerals at DSDP 612 by Glass (1988) provide a closer tie than previously available between tektites and impact craters.

We have previously examined the size and velocity distribution of solid ejecta fragments from impacts (O'Keefe and Ahrens, 1985, 1987). In this paper we examine the distribution of vapor condensate and liquid droplets for impacts on silicate and ice. We relate the former to the proposed Cretaceous/Tertiary impact event.

Specifically, we examine the issue of whether a sufficient mass of fine condensate could be produced by a large body impact such that it would reduce the solar insolation to the degree that a sudden worldwide decrease in temperature would occur. Finally, we examine the expected distribution of liquid droplets expected from such an event as a result of velocity fluctuations in the flow and compare these to observed tektite and micro-tektite size distributions.

IMPACT-INDUCED VAPORIZATION

The velocities required for the onset of vaporization of a variety of silicates and oxides that typify planetary materials were computed by Ahrens and O'Keefe (1972). For the impact of a silicate body on a silicate planet this onset occurs at a velocity in the range of 10-15 km/s depending upon the rock type. The mass of vapor and melt produced at these and higher velocities was calculated using equation of state data obtained from shock wave experiments, thermochemical data, and finite difference numerical techniques to solve the material response field equations (O'Keefe and Ahrens, 1977).

There are two vaporization regimes of interest; complete and partial. We define the complete vaporization regime as that in which the impact induced thermal energy is sufficient to completely vaporize the material (no liquid remains) upon expansion to 1 atmosphere pressure. Note that subsequently as the pressure approaches zero the vapor will condense into a liquid (e.g. Raizer, 1960). The partial vaporization regime is defined as that in which not all the material is vaporized upon release to 1 atmosphere and there is liquid entrained in the expanding vapor cloud. A schematic illustration of this is shown in figure 1 for a typical high velocity impact (V > 30 km/s). Referring to figure 1, the initial material that expands away from the impact site is highly shocked rock that is completely vaporized. This vapor will expand to a height many times the initial impactor radius before condensation will begin. The material below the vapor contains both vapor and entrained melt. This is defined as the partial vaporization regime. Since the impact thermal energy is not sufficient to completely vaporize the material, local pockets of vapor will be produced which will expand and accelerate the surrounding melt. The melt will be broken-up into small droplets by the velocity fluctuations in the expanding vapor. When the fraction of vapor to melt produced is small, then the expansion of the vapor to the point of condensation will produce frothy rock types such as found in the fall back ejecta of Barringer crater (Shoemaker, 1962). These rock types are found near the melt layer. The total mass of vapor produced, including that in a partial vaporization region, scales with the kinetic energy of the impactor at velocities above a vaporization threshold. The mass of vapor and melt as a function of impact velocity is shown in Fig. 2 (O'Keefe and Ahrens, 1977). The fraction of vapor produced in the complete vaporization regime at high velocities is 40%. At velocities less than 30 km/s all the vapor produced is in the partial vaporization regime. The amount of vapor produced at high velocities in the complete vaporization regime in terms of the equivalent radius (R_v) of a hemisphere of vapor at standard solid density as a function of impact velocity, can be readily obtained from figure 2. For the impact of a silicate projectile on a silicate planet,

$$R_v = 0.83 \ 10^{-4} R_i V^{2/3} \qquad V > 30\,km/s \tag{1a}$$

and for an ice projectile on an ice planet

$$R_v = 1.63 \ 10^{-4} R_i V^{2/3} \qquad V > 12km/s \tag{1b}$$

where R_i is the radius of the projectile (cm) and V is the velocity of impact (cm/s).

In the following sections of the paper we will first discuss the condensation of the vapor and distribution of particles in the complete vaporization regimes, and then the break-up and distribution of melt particles in the partial vaporization regime.

VAPOR SATURATION

Condensation of the vapor occurs because of the cooling produced by the isentropic expansion of the vapor. The onset of nucleation and growth of the condensates is initiated when the vapor cools to a point where it reaches saturation conditions. The conditions for saturation were determined by computing the intersection of the release isentrope of the vapor from the initial shock conditions and the vapor-tension curve (e.g. Raizer, 1960).

The average initial internal energy density of the vapor was determined from impact calculations of O'Keefe and Ahrens (1977).

The vapor-tension curve was assumed to have the following form

$$\rho = \frac{\exp(C)}{RT} \exp[-L/RT] \tag{2}$$

where T is the temperature and R is the universal gas constant, L is the molar heat of vaporization and C was adjusted to fit data for both silicates and water (Weast, 1982). These values are given in Table 1.

Table 1. Vapor tension curve values

	L(ergs/mol °K)	C	Temperature Range (K)
SiO_2	$3.63X10^{12}$	31.29	2005 to 2500
H_2O	$0.4255X10^{12}$	27.5	307 to 373

The impact induced vapor cloud was assumed to expand isentropically as a perfect gas. With this assumption, the mean density of the cloud as a function of the temperature is given by

$$\rho = \rho_i (T_i/T)^{1/(\gamma-1)} \tag{3}$$

where ρ_i and T_i are the initial density and temperature of the vapor and γ is the polytropic gas exponent.

The above equations can be solved for the density (ρ_{SAT})and temperature (T_{SAT}) of the vapor cloud at the onset of saturation. In addition, with the use of equation (1), the height above the point of impact where saturation occurs can be calculated. This height (h_s), as a function impactor radius for various impact velocities, is shown in Figure 3 for impacts of silicates on silicates and ice on ice. Note that this height is around 10 times the impactor radius and that for large scale impactors ($R_i > 5$ km) impacting the earth, the onset of saturation would occur well above the scale height of the atmosphere.

The mean velocity of expansion (V_{ex}) of the vapor cloud (Raizer, 1960) is given by

$$V_{ex} = (2\,E_m)^{1/2} \tag{4}$$

where E_m is the mean initial internal energy of the vapor, which is greater than the heat of vaporization (E_V) and was computed from code calculations of O'Keefe and Ahrens (1977). The internal energy relative to the heat of vaporization is given in Table 2.

The vapor mass was assumed to expand as a uniform hemispheric cloud (Raizer, 1960). In this case, the mean density (ρ) of the cloud as a function of time, (t), is given by

$$\rho = \rho_i (t_i/t)^3 \tag{5}$$

where

$$t_i \equiv R_i/V_{ex} \tag{6}$$

NUCLEATION

Under the dynamic conditions that occur during hypervelocity impact, condensation does not occur instantaneously when the cloud reaches the saturation point (Raizer, 1960). Nucleation of condensation centers starts at a very low rate at the time of saturation and reaches a maximum at some time later. The rate of nucleation is a very strong function of the degree of supersaturation, and to a sufficient degree of accuracy the total number of condensation centers or particles can be shown to be created at the time of maximum saturation (Raizer, 1960; Yamamoto and Hasegawa, 1977). We have used the theory of Yamamoto and Hasegawa (1977) and the extensions of Lattimer (1982) to compute the number of condensate particles and their radii under the conditions of meteorite impact. The relative time of maximum nucleation, (X_j), is given by the solution of

$$(X_j^3/a^2 + 1)^{-1}(6X_j^2/a^2 + 1)^{-1} = \frac{(\mu/\pi)^{1/2}}{108} \lambda^4 a^{-6} X_j^9 \exp(-a^2/X_j^2), \tag{7}$$

where

$$X_j \equiv (t - t_i)/\tau_{sat}$$

$$\tau_{sat} = \tau_c(b/kT_{sat} - 1)$$

$$a = \sqrt{4\mu^3/27}$$

$$\tau_c = -\frac{t_i}{3(\gamma - 1)}$$

$$\mu = 4\pi\sigma a^2/kT_{sat}$$

$$\lambda = \tau_{sat}/\tau_{coll}$$

$$\tau_{coll} = [\alpha C_1 4\pi a^2 V_T]^{-1}$$

where α is the sticking coefficient, σ the surface tension of the condensate, where $b = \Delta H/(R\text{-}1)$, and ΔH is the latest heat of formation of the solid, C_1 is the initial molecular concentration of the vapor, and V_T is the mean thermal velocity of the monomers, which is given by

$$V_T = \sqrt{kT_{sat}/2\pi m}$$

and m is the mass of the monomer.

The parameters in equation 7 are a, μ, and λ. The parameter, a, is a function of the molecular properties of the vapor and does not scale with the velocity of impact. The parameter, μ, is the ratio of the surface tension energy to the thermal energy. Because the saturation temperature varies by a factor which is only on the order of unity for variations in scales of impact from centimeters to hundreds of kilometers and velocity from 30 and 45 km/s, this parameter varies only a small amount. In contrast, λ varies significantly. This parameter is the ratio of the characteristic cooling time to the mean molecular collision time. The latter does not vary significantly with scale or velocity, however, the cooling time varies directly with the scale of the impact.

The total number of condensation centers created at the time of maximum nucleation determines the final radius of the condensate particles. The number of condensation centers produced after the time of maximum nucleation is negligible (Yamamoto and Hasegawa, 1977). The condensation centers grow until the condensate consumes all of the available vapor. The final radius, r, of the condensate spheres is given in dimensionless terms by

$$Z = \left(\frac{9}{8}\pi\right)^{1/6} X_j^2 a^{-4/3} \tag{8}$$

where the dimensionless radius is defined by

$$Z \equiv 3r[4\pi/(3\omega)]^{1/3}/\lambda$$

where ω is the molecular volume.

Shown in figure 4 is how variations in μ and λ influence the final radius of the condensate. Referring to figure 4, note that for typical variations in μ (e.g. factors of 2), the radius is not sensitive to μ, however, because λ can vary over many orders of magnitude as a result of the variation in cooling time with impact scale, this is a dominant parameter.

CONDENSATE MEAN RADII

Equations 1 through 8 form a sufficient set to calculate the mean radius of the condensates produced by impact. For a given impact velocity, equation 1 will give the radius of the equivalent hemisphere of vapor. The initial mean internal energy can be obtained from Table 2. Given the above, the temperature, the density at the point of onset of saturation can be calculated from the solutions of equations 2 and 3. Equation 5 can then be used to calculate the time of saturation. In turn, equation 7 can be used to compute the time of the maximum nucleation rate, then equation 8 is used to obtain the final radius of the condensate.

This procedure was carried out for the impact of silicate projectiles on a silicate planet and ice projectiles on an ice planet for impact velocities of 30 and 45 km/s and impactors varying from 1 to 10^7 cm in radius, using the values of the constants listed in Table 3. The height above the impact point at which the nucleation rate is maximum and the position where the condensation centers are formed is shown in Fig. 5. The final radius of the condensate particles as a function of the size of the impactor is shown in figure 6. The above calculations assume that there is not a planetary atmosphere or that the scale height is small as compared to the expansion scale of the vapor cloud. The primary effect of the atmosphere on the radius of the resulting condensates is through the retardation of the expansion of the vapor cloud by the atmosphere. This would change the cooling time of the vapor. We have made an assessment of this effect by parametrically reducing the expansion velocity. This effect on the condensate radius is shown in figure 7.

CONDENSATE PARTICLE SIZE DISTRIBUTIONS

In the above approach, we assumed that the density and internal energy are uniform throughout the vapor cloud in order to make the problem more tractable and to give a measure of the mean particle size. If the vapor cloud is uniform in its state properties then the distribution of particle sizes would be very monodispersed. The relative spread (Δr) in the particle sizes for a uniform vapor cloud due to the statistics of the condensation process is given by (Lattimer, 1982)

$$\Delta r/r_m = X_i\, a^{-2/3}\, (\pi/72)^{1/6} \tag{9}$$

where r_m is the mean radius. For typical impact condition values, the spread in particle size is less than 10 percent.

The dominant contribution to the range in particle sizes is the variation in the internal energy of the vapor cloud. A measure of this variation can be obtained for the results of detailed code calculations of O'Keefe and Ahrens (1977). The cumulative amount of vapor that expands with velocities greater than V, is given by

$$M_{CV}/M_{TV} = (V/V_{CV})^{-\delta}, \text{ for } V \geq V_{CV} \tag{10}$$

where M_{TV} is the total amount of vapor, that is in the complete vaporization regime, V_{CV} is the velocity at complete vaporization conditions, and δ fit to the impact calculations and is a function of impact velocity. The values of these curve fits are summarized in Table 2 along with the mean internal energy. In the case of condensation from the vapor, the high velocity case is appropriate and δ has a value in the range of 3.

The velocity of the vapor at complete vaporization conditions was calculated from equation 4.

$$V_{CV} = \sqrt{2E_v} \tag{11}$$

where E_V is the heat of vaporization.

From equation 10, the mass averaged mean internal energy can be obtained

$$E_m = E_{CV}\, (1/2)^{-2/\delta} \tag{12}$$

From these results, we see that the mean internal energy of the vapor increases with impact velocity. This is as would be expected, since the initial shock pressure and internal energy increase with velocity and the cessation of vaporization occurs under the same shock conditions independent of velocity.

For a given impact velocity and we can assess the effect on the condensate radius of a variation in the internal energy by constraining the expansion velocity to the mean value but varying the local internal energy around a condensate particle. The results of these calculations are shown in figure 8. The radius decreases with increasing internal energy but does not vary significantly with impact velocity. The results are tightly grouped except for deviations at small impact sizes ($r \cong 1$ cm). For particles greater than 10 cm

$$r/r_m = (E/E_{CV})^{-\alpha} \tag{13}$$

where $\alpha \cong 1.9$.

From the above information, the distribution of particle sizes can be derived. The amount of mass (dM) contained in particles whose radii are between r and r + dr is given by

$$dM = \frac{\partial M_{cv}}{\partial V}\frac{\partial V}{\partial E}\frac{\partial E}{\partial r}\,dr \tag{14}$$

Equation 14 can be evaluated using equations 4 and 10 through 13 with the result that the relative amount of mass is given by

$$\frac{dM}{M_{TV}} = \frac{\delta}{2\alpha}\left(\frac{r}{r_m}\right)^{\frac{\delta}{2\alpha}-1}\frac{dr}{r_m} \tag{15}$$

The number density is given by

$$dn = \frac{dM}{\rho_m \frac{4}{3}\pi r^3} = M_{TV}\,\frac{3\delta}{8\pi\alpha\rho_m r_m^3}\left(\frac{r}{r_m}\right)^{\left(\frac{\delta}{2\alpha}-4\right)}\frac{dr}{r_m} \tag{16}$$

where ρ_m is the density of the melt.

MELT DROPLET SIZE DISTRIBUTIONS IN THE PARTIAL VAPORIZATION REGIME

In the partial vaporization regime, the material which is ejected from the crater contains both melt and vapor. Since the shock induced thermal energy in a given volume is insufficient to completely vaporize the material, we expect that pockets of vapor will occur. The pockets of vapor are expected to be distributed randomly with some correlation with the orientation of defects such as cleavages and faults within the material. The vaporized regions will expand and accelerate the melt. Since the spatial distribution of the vapor pockets will not be uniform, velocity variations on the order of the expansion velocity are expected. These variations will result in the break-up of the melt into droplets. The approach to calculating the distribution of melt particles is similar to that of the condensate.

The amount of mass (dM_m) contained in melt particles whose radii are between r and r + dr is given by

$$dM_m = \frac{\partial M_{PV}}{\partial V} f(v) \frac{\partial V}{\partial r} dr \tag{18}$$

where $\frac{\partial M_{PV}}{\partial V}$ is the amount of both melt and vapor that is ejected at velocities between V as V + dV, and f(v) is the fraction of the total melt and vapor that is melt and $\frac{\partial V}{\partial r}$ is the relationship between ejection velocity and melt droplet radius.

The amount of melt and vapor can be obtained from equation 10, where in this case the velocity is less than or equal to the velocity of ejection required for complete vaporization

$$\frac{\partial M_{PV}}{\partial V} = -\frac{M_{TPV}}{V_{PV}\delta}\left(\frac{V}{V_{PV}}\right)^{-\delta-1} \qquad V \le V_{CV} \tag{19}$$

The values of δ in this case correspond to impact velocities less than 30 km/s and δ ranges from 5 to 4 for silicates.

The fraction (f_m) of mass ejected that is melt is given by

$$f_m(E) = \frac{E_{CV} - E}{E_{CV} - E_{PV}} \tag{20}$$

which when combined with equation 4 gives

$$f_m(V) = \frac{1 - \left(\frac{V}{V_{CV}}\right)^2}{1 - \left(\frac{V_{PV}}{V_{CV}}\right)^2} \qquad V_{PV} \leq V \leq V_{CV} \tag{21}$$

The maximum liquid particle size that is stable in a flow stream with a velocity difference ΔV between the particle and the fluid flow is again prescribed by the criteria that the Weber number is less than 10.

$$r = \frac{5\sigma}{\rho} \Delta V^{-2} \tag{22}$$

The density of the vapor ρ, is a function of the degree of vaporization and is given by

$$\rho = F \rho_o \equiv \frac{\left(\frac{E}{E_{PV}} - 1\right)}{\left(\frac{E_{CV}}{E_{PV}} - 1\right)} \rho_o \tag{23}$$

where ρ_o is the density of the melt.

The stable radius of the melt droplet when its energy approaches the energy of complete vaporization is

$$r_{CV} = \frac{5\sigma}{\rho_o f_V^2} (2 E_{CV})^{-1} \tag{24}$$

The stable radius of melt droplet when the energy approaches the incipient vaporization energy is

$$r_{PV} = \frac{5\sigma}{\rho_o f_V^2 \, F(E = 1.1 E_{PV})} (2 E_{PV})^{-1} \tag{25}$$

where F accounts for the formation of pumice and has a value in the range of 3.5×10^{-2}, the velocity difference, ΔV, is assumed to be some fraction, f_V, of the average expansion velocity V (see equation 9). Thus the relationship between the melt droplet radius and ejection velocity is given by

$$\frac{\partial V}{\partial r} = -\frac{1}{2f_V}\left(\frac{5\sigma}{\rho}\right)^{\frac{1}{2}} r^{-\frac{3}{2}} \qquad E > E_{PV} \tag{26}$$

Combining equations 18 through 23 gives for the mass distribution

$$dM = \frac{M_{TPV}}{2\delta}\left[\frac{\left(1-\frac{r_{CV}}{r}\right)}{\left(1-\frac{r_{CV}}{r_{PV}}\right)^{\frac{1}{2}}\left(1-\frac{r}{r_{PV}}\right)^{\frac{1}{2}}}\right]\left(\frac{r}{r_{PV}}\right)^{\frac{\delta}{2}-1}\frac{dr}{r_{PV}} \tag{27}$$

and for the number density distribution

$$dn = \frac{M_{TPV}}{\frac{8}{3}\pi\rho_m r_{PV}^3\delta}\left[\frac{\left(1-\frac{r_{CV}}{r}\right)}{\left(1-\frac{r_{CV}}{r_{PV}}\right)^{\frac{1}{2}}\left(1-\frac{r}{r_{PV}}\right)^{\frac{1}{2}}}\right]\left(\frac{r}{r_{PV}}\right)^{\frac{\delta}{2}-4}\frac{dr}{r_{PV}} \tag{28}$$

SUMMARY AND CONCLUSIONS

We have applied the homogeneous nucleation theory of Yamamoto and Hasegawa (1977) and its extensions by Lattimer (1982) to calculate the radii of condensate particles produced in planetary impact. The scope of applications range from small scale (~ 1 cm) impacts to those representative of collisions of proto-planetesimals (~10^7 cm) for a range of impact velocities from 15 to 45 km/s. We considered the cases of silicate impactors on silicate planets and ice impactors on ice planets.

At impact velocities above 30 km/s significant quantities of vapor are produced. This vapor expands in a nearly hemispherical cloud away from the point of impact. The vapor cools during the expansion and reaches a point where nucleation of condensation centers is a maximum and particulate growth is initiated. The height above the impact point at which maximum nucleation occurs is on the order of ~10 times the impactor radii for silicate and ice impacts (see figure 5). As an example, consider the case of an impact representative of the K-T event (Alvarez et al., 1980) which is assumed to have an impactor of radius of 5 km. The height above the Earth at where condensation would be initiated is greater than 35 km, which is greater than the 7 km scale height of the atmosphere. This implies that the condensates would be formed outside of the Earth's atmosphere. The condensates would solidify outside the atmosphere and would be propelled on ballistic trajectories and be distributed globally. The particulates would decelerate upon reentry into the atmosphere and the larger particles would be expected to have ablation features found in tektites. The mean velocity of expansion of the vapor, not accounting for any reduction due to the atmosphere, is close but does not exceed the Earth's escape velocity; thus allowing for the reentry of the particles. The mean velocity of expansion of the silicate vapor is ~6 km/s. This value exceeds the escape velocity of most of the planets and satellites, except for Earth and Venus.

The mean condensate radii were calculated and the results shown in figure 6. The mean size is directly proportional to the scale of the impact for a given impact velocity. The effect of impact velocity has competing processes. Increasing the velocity of impact, increases the amount of vapor produced. If the mean internal energy did not change with velocity, then the size of the condensate would proportional to the effective increase of the radius of the vapor sphere. However, the mean internal energy does increase with impact velocity (see Table 2). This tends to decrease the size of the condensate. Because of these compensating effects, an increase in impact velocity of a factor of 3 results in a condensate radii increase of only 30%.

The primary effect of a planetary atmosphere on the condensate radii is through the retardation of the expansion of the vapor cloud. This effect is important when the atmospheric scale height is greater than the nucleation height. The effect was assessed parametrically by reducing the mean velocity of expansion. Referring to figure 7, we see that by reducing the mean velocity of expansion by a factor of two, this increases the mean particle size by that amount.

If the internal energy distribution in the vapor cloud was uniform then the variation in condensate particle sizes for a given event would be small (~10%). However, in the case of hypervelocity impact, the initial shock wave that is produced decays with distance and results in a variation in the internal energy. We have assessed this effect on the particle distribution and found that it considerably broadens it. Referring to equation 15, we see that the relative amount of mass in a relative size increment varies as $r^{-\delta/2\alpha}$. For typical values of these parameters for silicates, the exponent ranges from -1.05 to -0.84. This implies that the relative amount of mass in a given particle range varies inversely with the radius of the particle. The number distribution is given by equation 16. For the range of values of the parameters (δ, α), the associated slope varies from -2.05 to -1.84. B. Glass (personal communication, 1972) measured the number distributions for microtektites in the Australasian and Ivory Coast fields (Table 4).

At velocities above 10 km/s significant quantities of material are partially vaporized. The pockets of vapor produced break-up the melt into droplets. The mass and number distributions were calculated and are given in equations 27 and 28. Number distribution was compared with the tektite distribution measured by B. Glass (1972). The model distribution shows good agreement with the measured distribution (see figure 9). The range of particle sizes is relatively narrow and extends only over about an order of magnitude in radii from 500 to 5000 microns.

In the case of the K-T impact event, the mean condensate radius is in the range of several centimeters. The important consideration for biological extinction is the amount of mass in the sub-micron sized particles because these have long atmospheric residence times ($\sim 10^7 - 10^8$ sec). The relative amount of mass in micron-sized particles can be calculated from equation 15. The relative amount of mass is 10^{-4} that of the impactor. The amount of vapor produced would be in the range of 10^{18} to 10^{19} g. This implies that 10^{14} to 10^{15} g of sub-micron sized particles could be produced. Gerstl and Zardecki (1981) require around 10^{16} g distributed worldwide to cause significant solar extinction. Hence the calculated mass of fines produced via condensation at very high altitudes is a $\sim 10^{-1}$ to 10^{-2} of the minimum value estimated to be required to affect world-wide climate. The particle sizes that are expected from break-up of melt in the partial vaporization regime are too large ($500 < r < 5000$ microns) to have the required long atmospheric residence times. O'Keefe and Ahrens (1987) have calculated the mass distribution function for the dust particles, and within the uncertainties in the empirical database on small scale impacts, concluded that the amount of dust would be an order of magnitude greater than the values given above, and probably sufficient to significantly decrease solar insolation for the impact of a 10 km diameter bolide.

The approach we have taken relies on the use of the homogeneous nucleation theory of Yamamoto and Hasegawa, 1977. Condensation silicates have been an active area of research for decades, and there is still much work to be done because of the complexity of the phenomena.

The conclusions using the condensation model are that for impactors of diameters of 10 km or less, the condensates from the impact vapor will be too large to have long atmospheric residence times and thus would <u>not</u> cause significant extended decreases in solar radiation.

Acknowledgments

Research supported under NASA Grant NSG-7129. We are grateful to B. Glass for the data on microtektite distributions (figure 9). Contribution number 4660, Division of the Geological and Planetary Sciences, California Institute of Technology, Pasadena, California.

REFERENCES

Ahrens, T. J. and J. D. O'Keefe (1972) Shock melting and vaporization of lunar rocks and minerals, *The Moon, 4, no. 3*, 214-249.

Alvarez, L. E., Alvarez, W., Asaro, F., and Michel, H. V., Extraterrestrial cause for the Cretaceous-Tertiary extinction, *Science, 208*, (1980), 1095-1108.

Courtillot, V, J. Besse, D. Vandamme, R. Montigny, J. J., Jaeger and H. Cappetta, 1986, Deccan flood basalts at the Cretaceous-Tertiary boundary? *Earth & Planet. Sci. Lett., 80,* 361.

Gerstl, S. A. W., and A. Zardecki, Reduction of photosynthetically active radiation under extreme stratospheric aerosol loading: Proc. Conf. Large Body Impact, Sp. Paper 190, *Geol. Soc. Am.*, 201-210, 1982.

Lattimer, J. M., 1982, Condensation of grains, in *Formation of Planetary Systems*, ed by. A. Brahic, Centre Natationeil D'Etudes Spatiales, Toulouse, 189-282.

J. M. Luck and K. K. Turekian, 1982, Osmium - 187/Osmium - 186 in Manganese nodules and the Cretaceous-Tertiary boundary, *Science 222*, 613-615.

Officer, C. B., and C. L. Drake, 1985, Terminal Cretaceous environmental events, *Science 227*, 1161.

Officer, C. B., and C. L. Drake, 1983, The Cretaceous-Tertiary Transition, *Science 219*, 1383.

O'Keefe, J. D. and T. J. Ahrens (1977) Impact-induced energy partitioning, melting, and vaporization of terrestrial planets, *Proceedings, Lunar Science Conf., (1977a) 8th*, 3357-3379.

Raizer, Y. P., Movement of a gas under the action of a concentrated impact on it's surface (during an explosion on the surface), *Zhurnal Prikladnoy Mekhaniki i Tekhnicheskoy Fiziki, Nr. 1*, 57-66 (Foreign Tech. Div. Transl. 63-80/AD 434 954), 1963.

Raizer, Y. P., 1960, Condensation of a cloud of vaporized matter expanding in vacuum: Sov. Phys.-JETP, v. 37, p. 1229-1235.

Scarlett, B. and Buxton, R. E., 1974, Particle size distribution of spherical particles in Apollo 12 samples: Earth and Planet. Sci. Lett., v. 22, p. 177-187.

Weast, R. C., 1982, *Handbook of Chemistry and Physics* 63rd Ed. PD 219-220, CRC Press. Boca Raton.

Yamamoto, T. and H. Hasegawa, Grain formation through nucleation process in astrophysical environments, *Prog. Theor. Phy., 58*, 816-828.

Table 2. Variations of mean velocity of expansion and thermal energy as a function of impact velocity (from O'Keefe and Ahrens, 1977)

IMPACT VELOCITY (KM/S)	δ	$\frac{V}{V_{cv}}$	$\frac{E_m}{E_v}$
15	5	1.15	1.31
30	4	1.18	1.41
45	3.2	1.24	1.54

Table 3. Constraints used in condensation calculations

PARAMETER	SILICATE	ICE
σ (ergs/g)	875	7.0
γ	1.4	1.3
ω (cm^3)	1.43×10^{24}	1.4×10^{24}
Energy for incipient vaporization, E_{iv} (ergs/gm)	0.05×10^{10}	2.36×10^{10}
Energy of complete vaporization, E_v, (ergs/gm)	1.8×10^{11}	2.9×10^{10}

Table 4. Number distribution of microtektite*

	Microtektites (% abundance)	
diam (μm)	Australasian	Ivory Coast
125-175	47	41
176-245	30	30
246-350	14	19
351-495	6	9
496-707	2	1
708-1000	0.5	0.5
Total Number	264	621

* Unpublished data from paper presented at spring meeting of the American Geophysical Union, 1972, by B. Glass

Figure Captions

Fig. 1 Schematic illustration of the states of the ejecta material.

Fig. 2 Relative amount of melt and vapor produced as a function of dimensionless impact velocity (O'Keefe and Ahrens, 1977), ρ_2 and ρ_1 are the densities of the impactor and planetary surface, V is the impact velocity and $C_p \equiv 7.44 \times 10^5$ cm/s. CV and PV are the relative amounts of material completely and partially vaporized.

Fig. 3 Height above the point of impact at which saturation occurs. Results are plotted for impact velocities of 30 and 45 km/s for impacts of silicate on a silicate planet and ice on an ice planet.

Fig. 4 Radius of the condensate particle as a function of the ratio of saturation time/collision (λ) time for various values of the ratio of surface tension energy/thermal energy (μ).

Fig. 5 Height above the point of impact at which the nucleation rate is a maximum. Results are plotted for impact velocities of 30 and 45 km/s for impact of a silicate projectile on a silicate planet and an ice projectile on an ice planet.

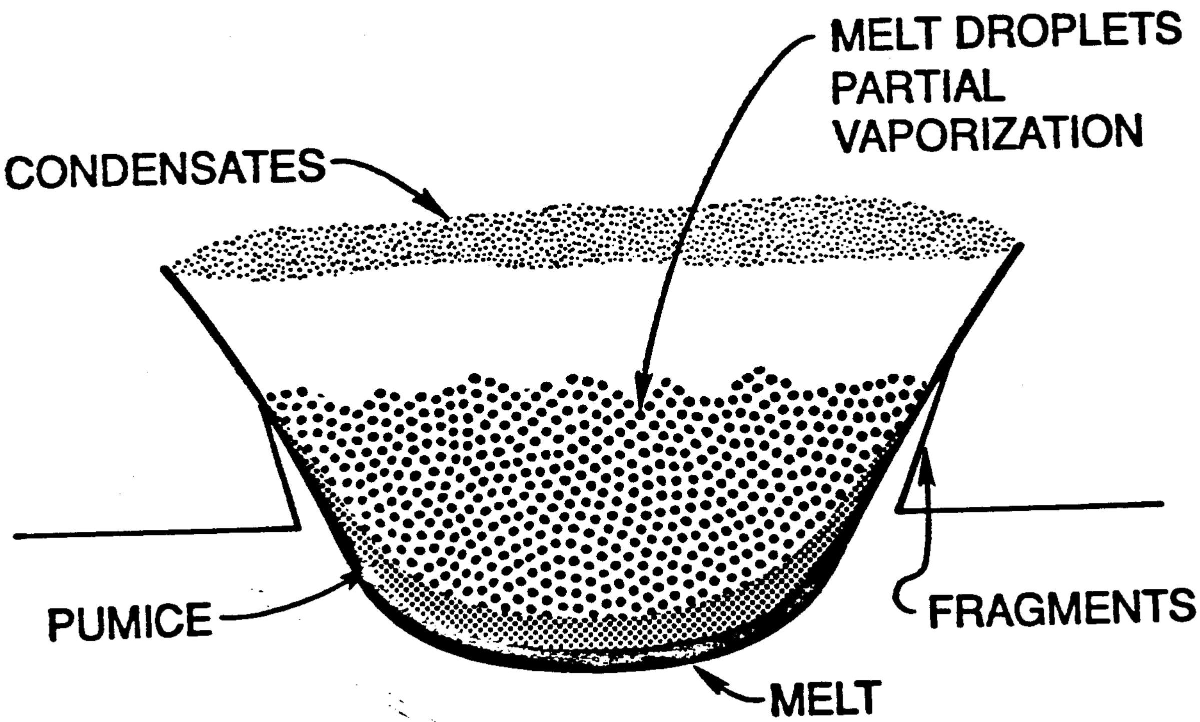
Fig. 6 Condensate radius as a function of impactor radius. Result as plotted for impact velocities of 30 and 45 km/s for impact of a silicate projectile on a silicate planet and an ice projectile on an ice planet.

Fig. 7 Condensate radius as a function of compaction radius for cases in which the expansion rate was varied by factors of 1, 0.5, and 0.1 The reduction in expansion velocity simulates the effect of a planetary atmosphere in retarding its expansion.

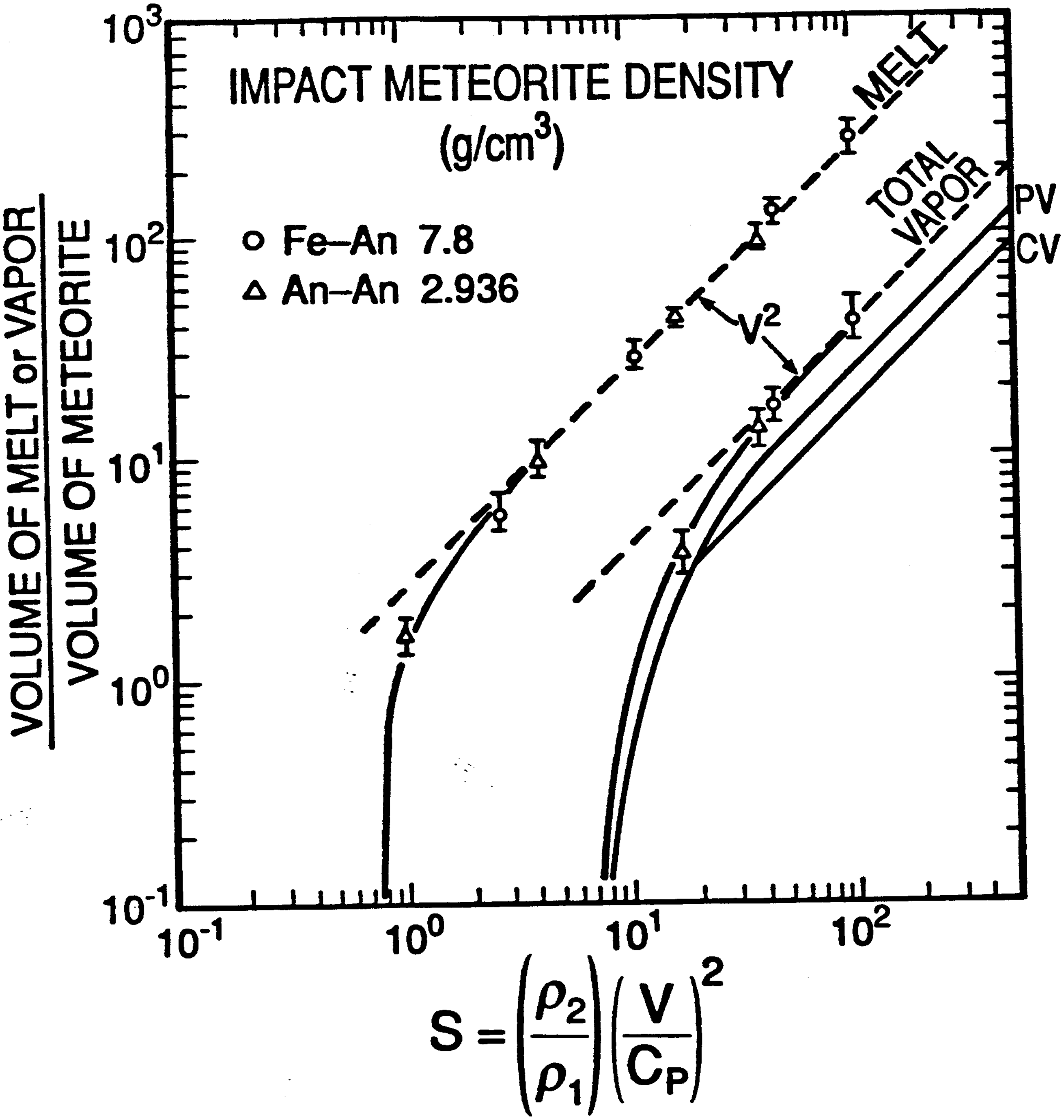
Fig. 8 The condensate radius divided by the mean condensate radius as a function of variation of the local internal energy from the mean for a 1 km, 45 km/sec, silicate impactor. The expansion velocity was held constant.

Fig. 9 The relative abundances of microtektite (B. Glass, 1972) solid line is the calculated abundance using the partial vaporization model (equation 28).

EJECTA MATERIAL STATES

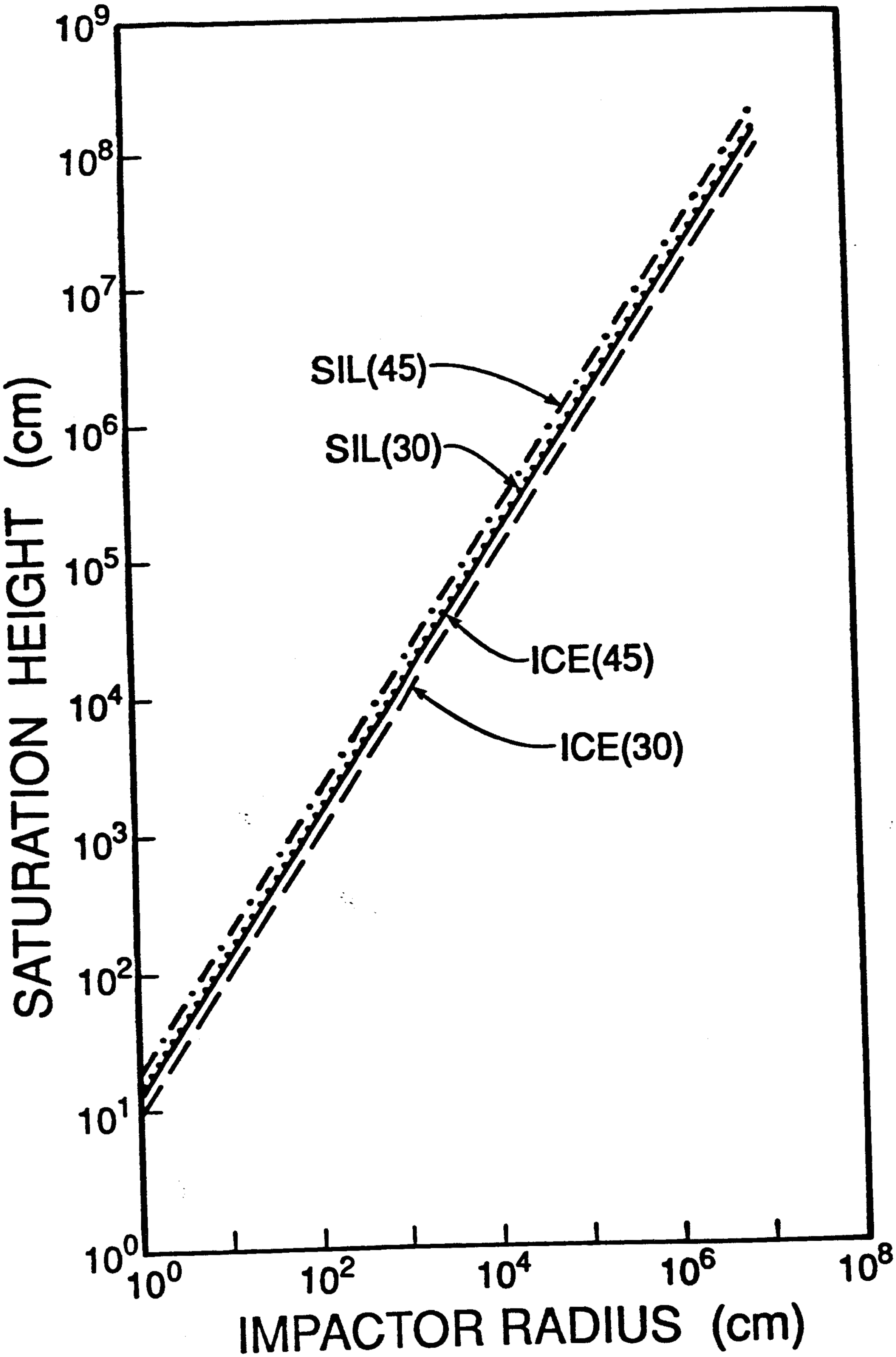


TJA89032SFD

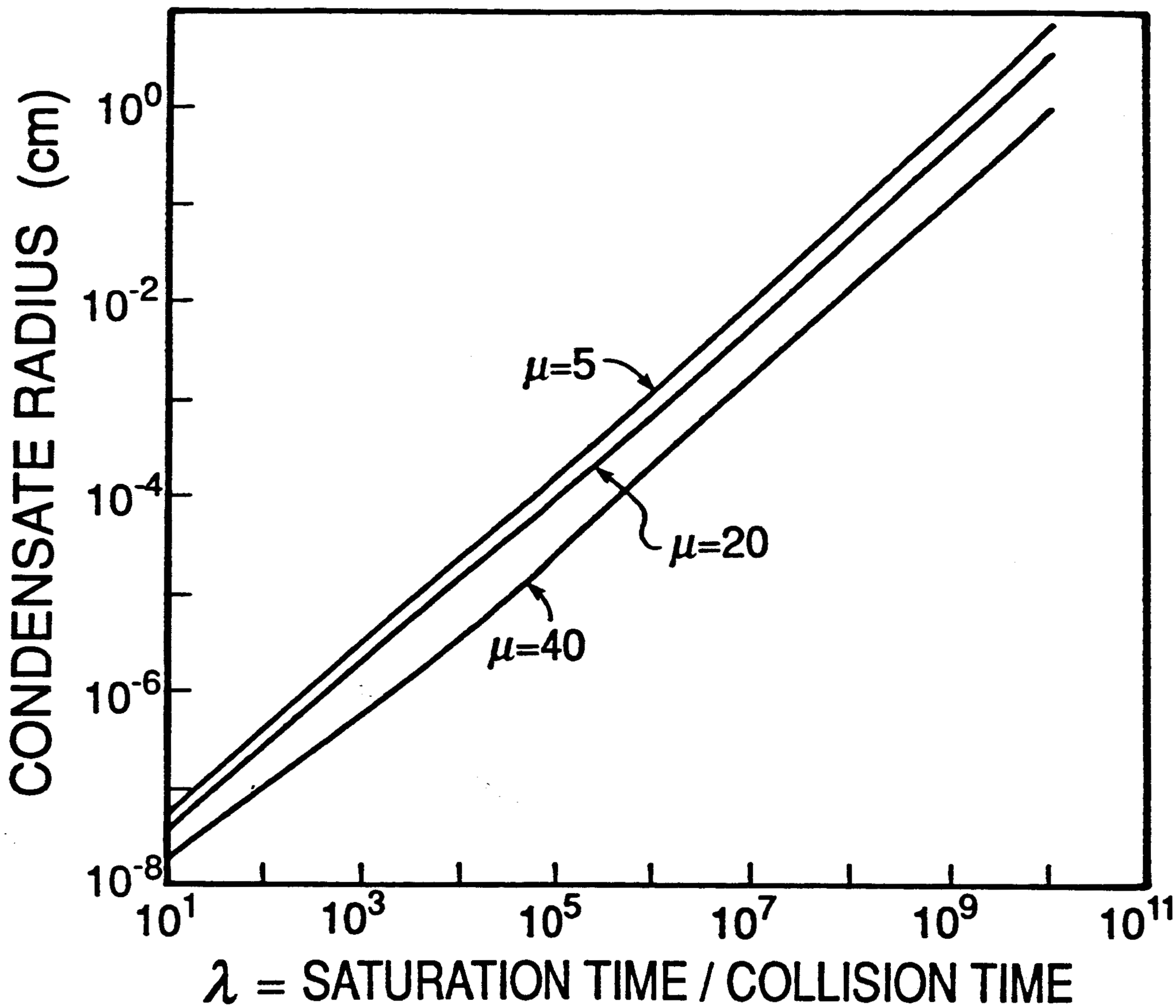


TJA89033SFD

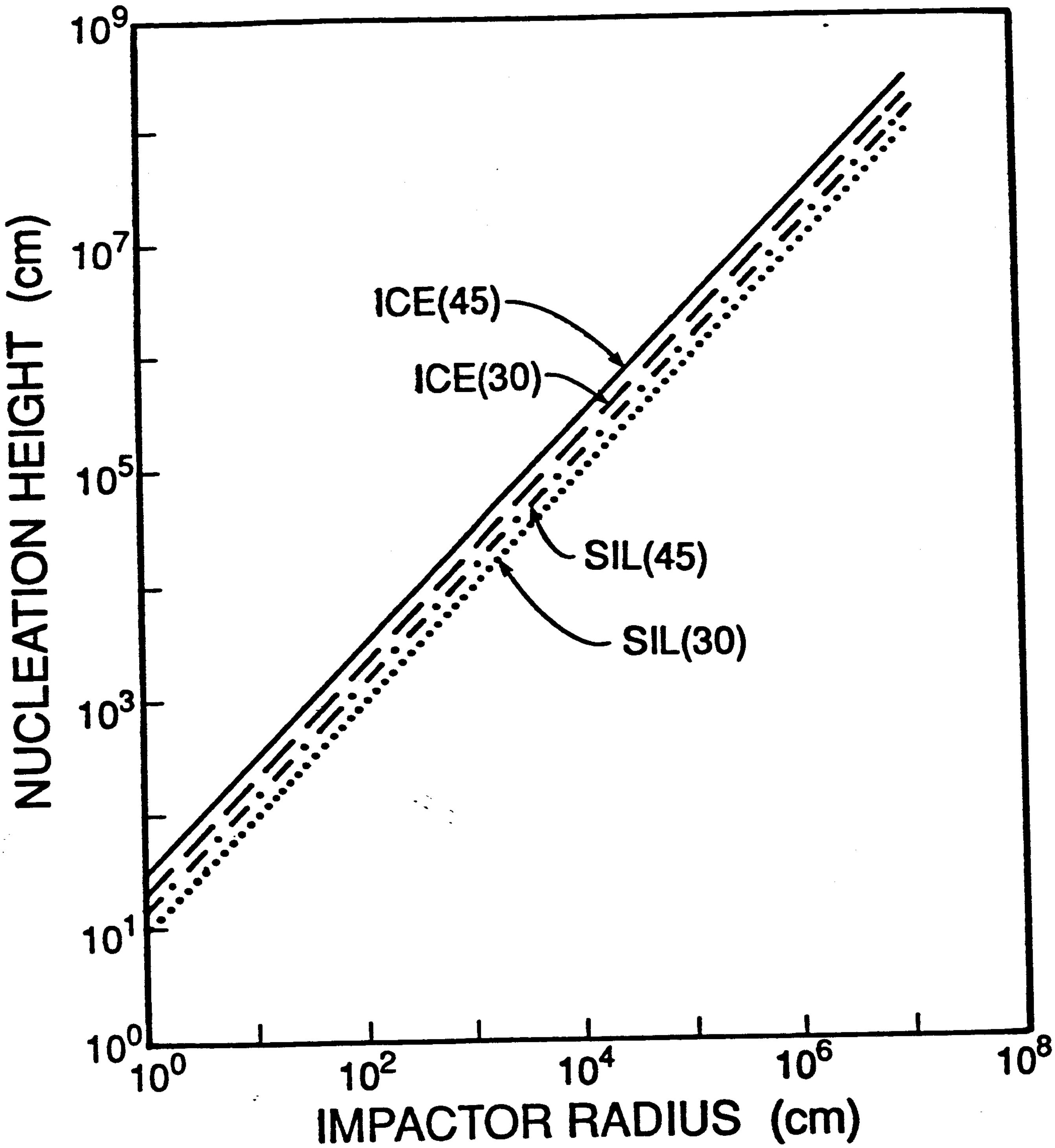
2



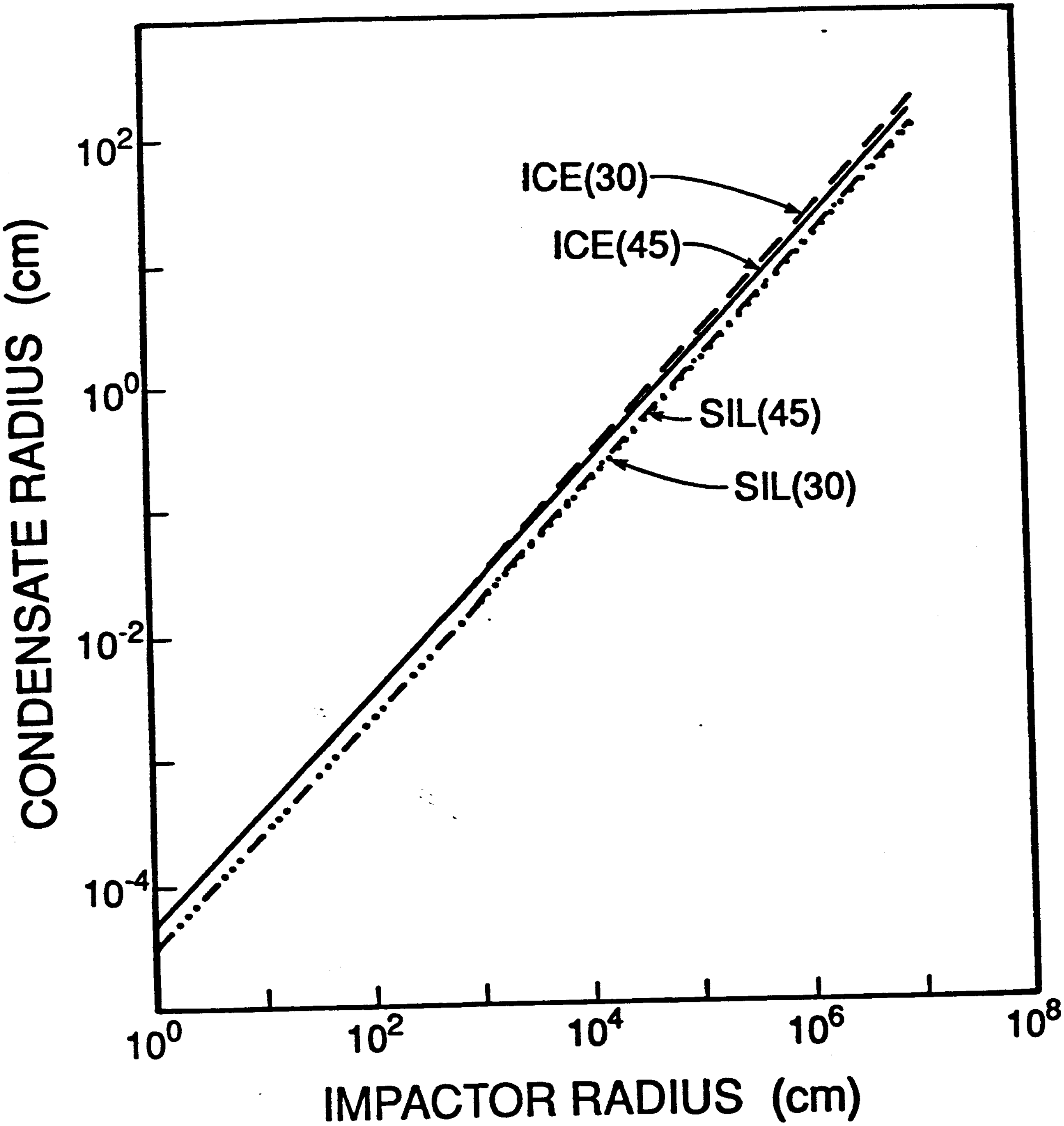
TJA89034SFD



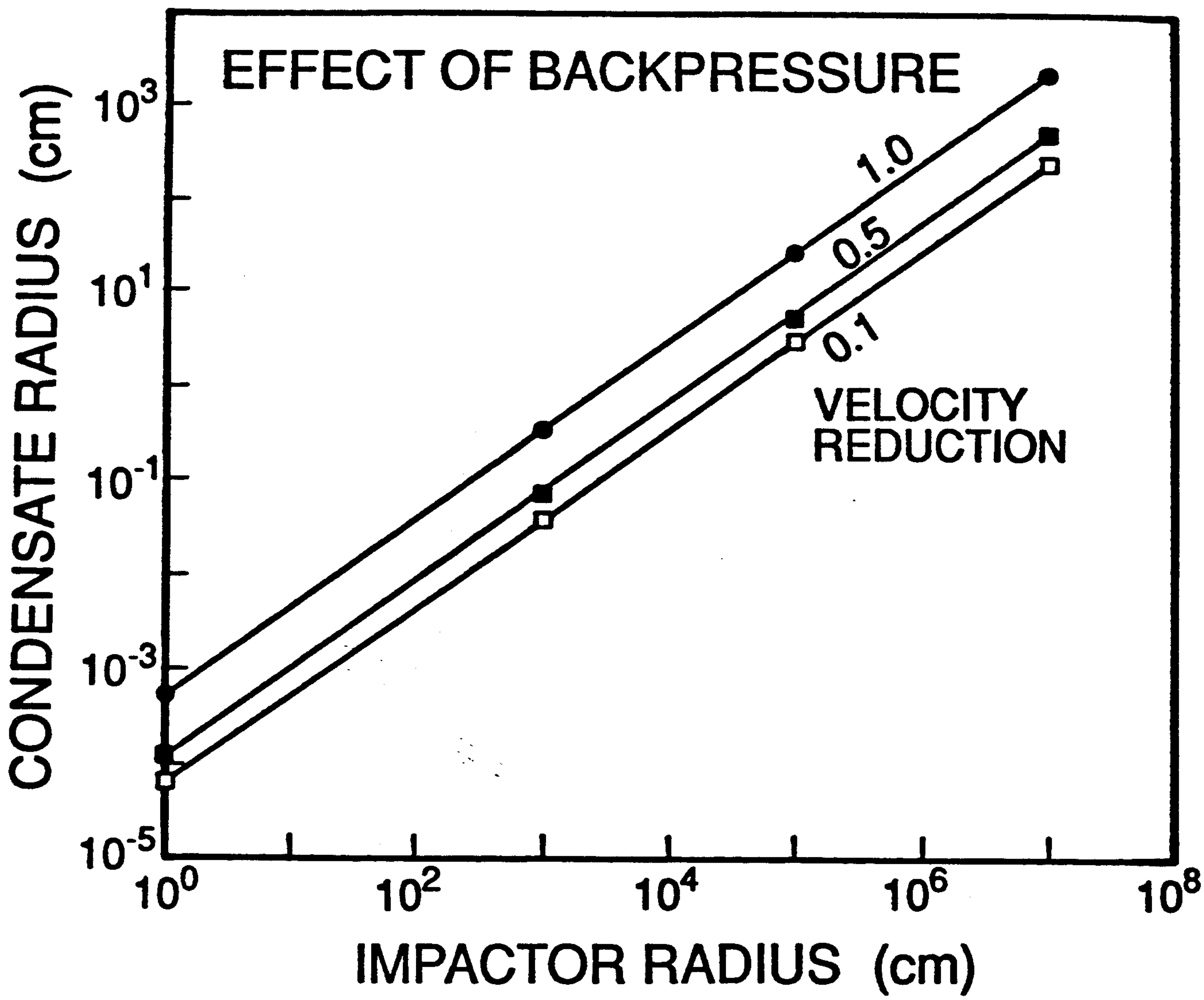
CONDENSATE RADIUS (cm)
10^0
10^{-2}
10^{-4}
10^{-6}
10^{-8}
10^1
10^3
10^5
10^7
10^9
10^{11}
μ=5
μ=20
μ=40
λ = SATURATION TIME / COLLISION TIME

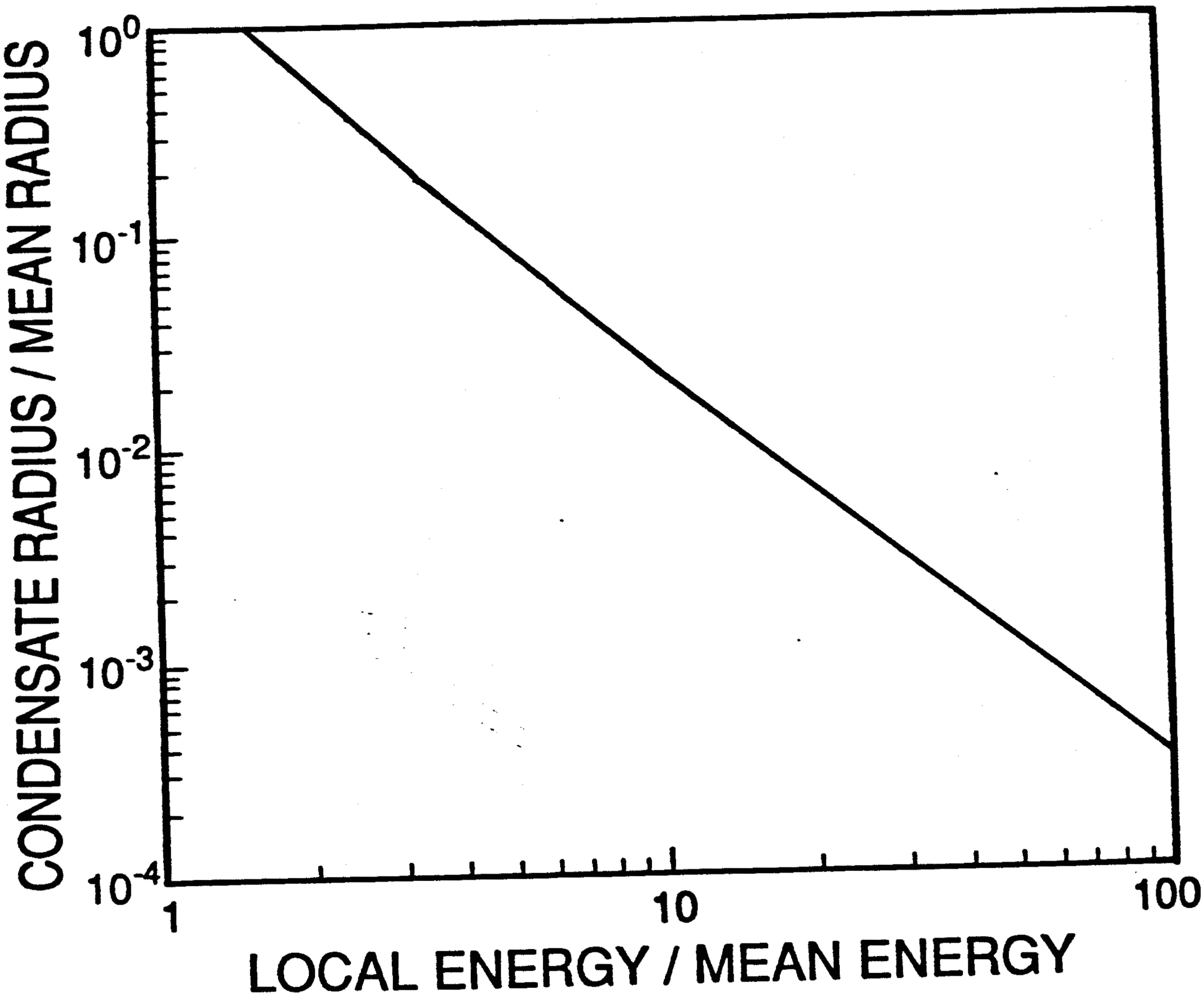


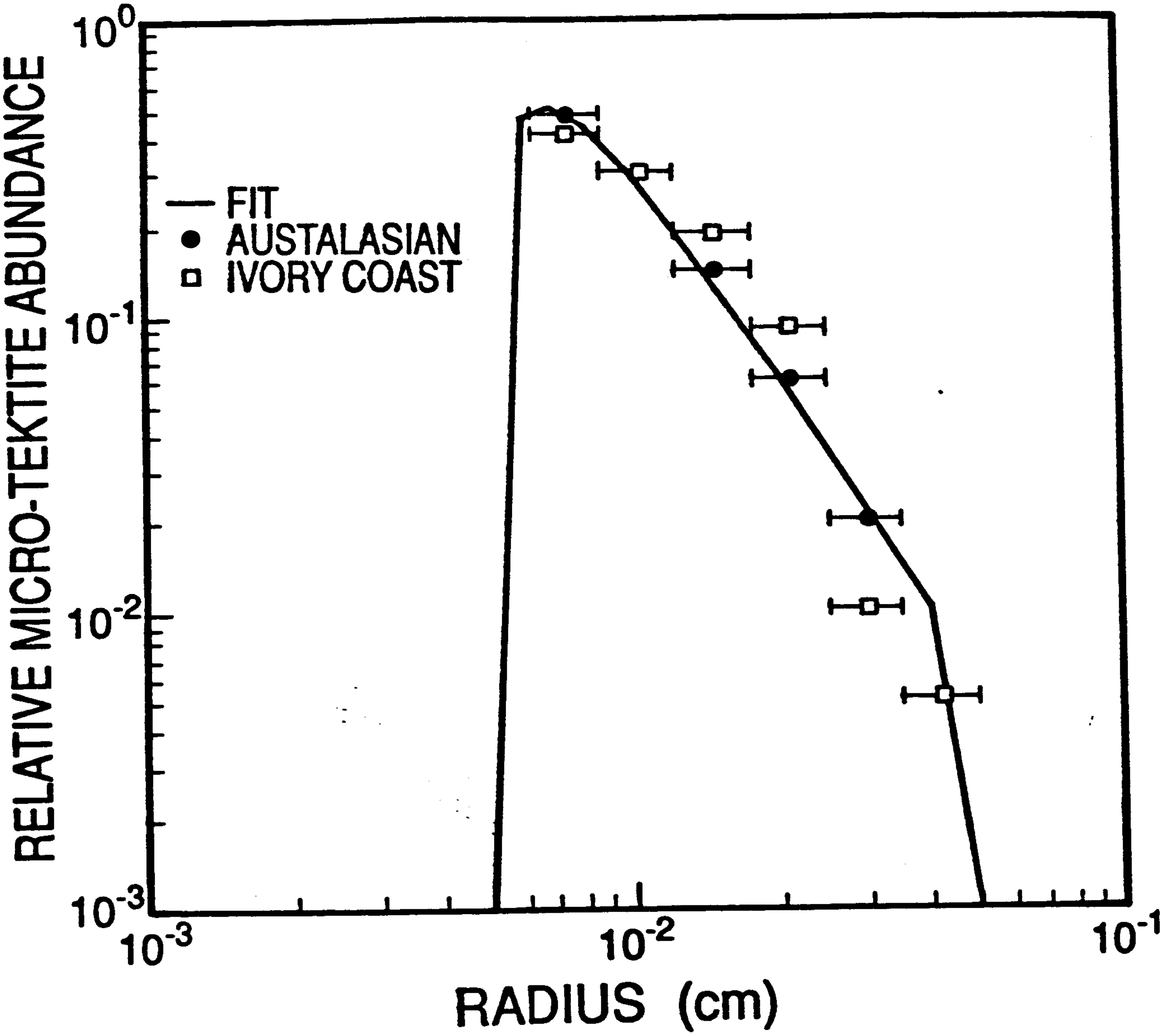
TJA89036SFD



TJA89037SFD


EFFECT OF BACKPRESSURE
CONDENSATE RADIUS (cm)
IMPACTOR RADIUS (cm)
1.0
0.5
0.1
VELOCITY REDUCTION
10^{3}
10^{1}
10^{-1}
10^{-3}
10^{-5}
10^{0}
10^{2}
10^{4}
10^{6}
10^{8}


CONDENSATE RADIUS / MEAN RADIUS
10^0
10^{-1}
10^{-2}
10^{-3}
10^{-4}
1
10
100
LOCAL ENERGY / MEAN ENERGY



RELATIVE MICRO-TEKTITE ABUNDANCE
10^0
10^{-1}
10^{-2}
10^{-3}
— FIT
● AUSTALASIAN
□ IVORY COAST
10^{-3}
10^{-2}
10^{-1}
RADIUS (cm)